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SNOW CHARACTERISTICS IN DRONNING MAUD LAND, ANTARCTICA

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SNOW CHARACTERISTICS IN DRONNING MAUD LAND, ANTARCTICA

Eija Kanto (née Kärkäs)

ACADEMIC DISSERTATION IN GEOPHYSICS

To be presented, with the permission of the Faculty of Science of the University of Helsinki for public criticism in the Auditorium D101 of Physicum, Gustaf Hällströmin katu 2, on September 15th, 2006, at 12 o'clock noon.

Helsinki 2006

We came to probe the Antarctic's mystery, to reduce this land in terms of science, but there is always the indefinable which holds aloof yet which rivets our souls.

Sir Douglas Mawson, The Home of the Blizzard

Antarctica left a restless longing in my heart beckoning towards an incomprehensible perfection forever beyond the reach of mortal man. Its overwhelming beauty touches one so deeply that it is like a wound.

Edwin Mickleburgh, Beyond the Frozen Sea

Contents

This thesis is based on the following five articles, which are referred to in the text by their Roman numerals:

I Kärkäs, E., H.B. Granberg, C. Lavoie, K. Kanto, K. Rasmus and M. Leppäranta. 2002. Physical properties of the seasonal snow cover in Dronning Maud Land, East-Antarctica. *Annals of Glaciology* **34**, 89-94.

II Kärkäs, E., T. Martma and E. Sonninen. 2005. Surface snow properties and stratigraphy during the austral summer in western Dronning Maud Land, Antarctica. *Polar Research* **24**(1-2), 55-67.

III Kärkäs, E., K. Teinilä, A. Virkkula and M. Aurela. 2005. Spatial variations of surface snow chemistry during two austral summers in western Dronning Maud Land, Antarctica. *Atmospheric Environment* **39**, 1405-1416.

IV Kanto, E., K. Teinilä, H. Timonen and E. Sonninen. Stratigraphy and spatial variations of snow chemistry in western Dronning Maud Land, Antarctica. Submitted to *Nordic Hydrology*.

V Kärkäs, E. 2004. Meteorological conditions of the Basen nunatak in western Dronning Maud Land, Antarctica, during the years 1989-2001. *Geophysica* **40**(1-2), 39-52.

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Abstract

Snow cover is very sensitive to climate change and has a large feedback effect on the climate system due to the high albedo. Snow covers almost all surfaces in Antarctica and small changes in snow properties can mean large changes in absorbed radiation. In the ongoing discussion of climatic change, the mass balance of Antarctica has received increasing focus during recent decades, since its reaction to global warming strongly influences sea-level change.

The aim of the present work was to examine the spatial and temporal variations in the physical and chemical characteristics of surface snow and annual accumulation rates in western Dronning Maud Land, Antarctica. The data were collected along a 350-km-long transect from the coast to the plateau during the years 1999-2004 as a part of the Finnish Antarctic Research Programme (FINNARP). The research focused on the most recent annual accumulation in the coastal area.

The results show that the distance from the sea, and the moisture source, was the most predominant factor controlling the variations in both physical (conductivity, grain size, $\delta^{18}O$ ratio and accumulation) and chemical snow properties. The sea-salt and sulphur-containing components predominated in the coastal region. The local influences of nunataks and topographic highs were also visible on snow. The variations in all measured properties were wide within single sites mostly due to redistribution by winds and sastrugi topography, which reveals the importance of the spatially representative measurements. The mean accumulations occurred on the ice shelf, in the coastal region and on the plateau: 312 ± 28, 215 ± 43 and 92 ± 25 mm w.e., respectively. Depth hoar layers were usually found under the thin ice crust and were associated with a low dielectric constant and high concentrations of nitrate. Taking into account the vast size of the Antarctic ice sheet and its geographic characteristics, it is important to extend investigation of the distribution of surface snow properties and accumulation to provide well-documented data.

Acknowledgements

There are several people and groups I like and should thank and without whom this thesis would have never been completed:

My supervisor Professor Matti Leppäranta gave me the privilege of working in the Antarctic project. He has given me a lot of independence, responsibilities and guidance during this journey, as well as my other supervisor Professor Hardy Granberg, who introduced me to snow business in the first place. They are both gratefully acknowledged.

Doc. Aki Virkkula and Dr. Kimmo Teinilä led me into the field of snow chemistry. They have both shown tireless supervision and the true meaning of collaboration.

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My husband Kimmo stood beside me during this work and he is the one who understands without any doubts my love of the polar regions. I want to thank my closest relatives for encouraging but never questioning my choices. And my friends have sometimes found my topic even more fascinating than I have.

1. Introduction

Polar regions (Fig. 1), dominated by ice sheets and sea ice, are universally recognized as extremely important for cooling the climate of our planet (Bindschadler, 1998). The hemispherical reflectance of ice sheets can exceed 90%, making them the brightest surfaces on Earth. The most obvious climatic effect of ice sheets resides in their influence on the earth's heat budget through changes in the planetary albedo (i.e. ratio of reflected to incident solar radiation) (Muszynski and Birchfield, 1985). Their low temperatures, combined with the heat of the tropics, create a temperature gradient that contributes to the meridional exchange of heat by atmospheric circulation (Bindschadler, 1998). Ice sheets are the largest freshwater reservoirs on Earth and constitute a unique archive of past climate and environmental changes (e.g. Petit et al., 1999).

Snow and ice cover make the Antarctic continent one of the principal components of our global climate system. Snow covers 98% of all surfaces in Antarctica and has a major effect on the continent's cold climate due to its high albedo. Precipitated snow also represents the input into the mass balance of the continent. The changes in mass balance are important but difficult to quantify. They influence sea levels and currents worldwide. The estimates place 70-80% of the world's freshwater in ice sheets and glaciers and Antarctica is, by far, the largest component with 89% (~$30*10^6$ km^3; Delaygue et al., 2000) of this frozen fraction (Bindschadler, 1998). In view of the predicted changes in mass balance of Antarctic snow, we need to understand the detailed spatial variations in it and its annual accumulation much better than we currently do.

The properties of surface snow give the continent its important climatic role and determine its energy and mass balance. To better understand these relationships it is essential to understand the properties of snow with their spatial variations. Since snow covers nearly everything in Antarctica, it is the snow cover that generates the signal detected by a variety of remote sensing satellites. To better understand these signals, it is essential to know the physical properties of the snow cover that influence its absorption, backscatter and emission properties. The surface properties of snow are valuable ground-truth data for satellite remote sensing and large-scale numerical modelling, which allow us to study large and otherwise inaccessible areas (König et al., 2001).

Antarctica provides the cleanest atmospheric environment available for study of the chemicals stored in snow and accumulated on the polar ice sheet, although the atmosphere of the polar regions is already affected by human activities (Legrand and Mayewski, 1997). Ice cores give access to palaeoclimate records that include local temperature and precipitation rate, moisture source conditions, wind strength and aerosol fluxes of marine, volcanic, terrestrial, cosmogenic and anthropogenic origin (Petit et al., 1999). Understanding the spatial variability of ion concentrations in the snow cover is essential for interpreting the climate records from deep ice cores.

The physical and chemical properties of the snow cover interact with each other, as well as with the meteorological conditions that influence this complex system. The meteorological conditions of Antarctica are characterized by pronounced seasonal cycles with a long polar night, very cold temperatures, strong winds and dryness of the air (King and Turner, 1997).

In 1999 the project 'Seasonal snow in Antarctica' was initiated and was funded by the Academy of Finland in two phases: 1999-2001 and 2002-2005. The project resulted from collaboration between the Division of Geophysics from the University of Helsinki and the Centre d'Applications et de Recherches en Télédétection (CARTEL), Université de Sherbrooke, Québec, Canada. The general objective was to increase the level of understanding of the surface snow cover in Dronning Maud Land, Antarctica, and its role in the Antarctic as well as global climate. The project's strategy was to develop detailed knowledge of the physical properties of snow, solar radiation of the snow surface, snow temperature evolution during the Antarctic winter and further for generating these properties to use remote sensing and modelling and to expand this knowledge to Antarctica as a whole.

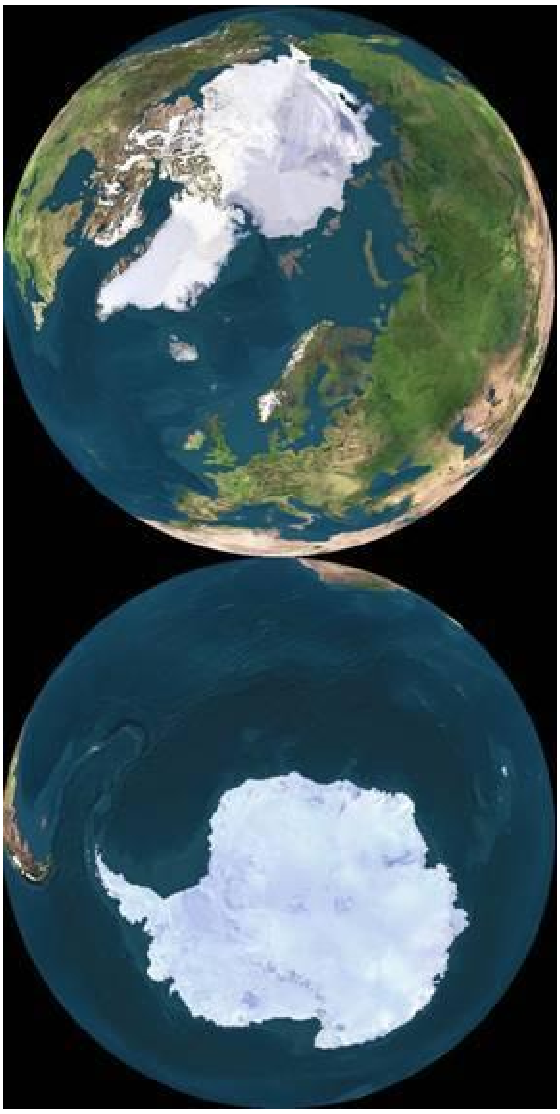

Figure 1. Snow- and ice-covered polar areas seen from space during the corresponding summer (Satellite data provided by The Living Earth® Inc. /Earth Imaging © 2005).

The present study concerns the physical and chemical properties of the surface snow cover in the western Dronning Maud Land area and is a part of the ‘Seasonal snow in Antarctica’ project. The primary objective was to observe the spatial and temporal variations in surface snow properties and annual accumulation in the measurement area. The aim was to determine the properties of surface snow in the area and examine what environmental factors produce them and provide a data that could be used in remote sensing as well as in interpreting the ice cores.

The physical properties of snow and its annual accumulation rates (I, II) constitute the leading glaciological foundations of this work. During the time span of this investigation knowledge of the associated problems increased and it seemed important to expand the study to include the chemical properties (III, IV). Snow properties are dependent on the prevailing meteorological conditions in the area. A general overview of these meteorological conditions is outlined (V).

2. Glaciochemical properties of snow cover

2.1 Snow formation and metamorphosis

Formation of snow in the atmosphere begins when the temperature is less than 0 °C and supercooled waters and foreign nucleation sites (e.g. sea-salt, dust and mineral particles) are present. Water vapour condenses around nuclei. Once an ice crystal is nucleated, it grows through aggregation of small ice crystals and riming from water droplets into the snowflake form, and finally lands on the ground as snowfall. The temperature at which it grows determines the basic shape of an ice crystal. The rate of growth and secondary crystal features are determined by the degree of supersaturation. The falling snow forms a porous cover consisting of ice crystals, with liquid water and moist air in its pores. By the time snow reaches the ground it has already undergone a number of transformations. After the snow is deposited thermodynamical processes known as metamorphism modify the particle shapes. Changes in snow properties are driven by the energy, mass and momentum exchange in the surface and bottom boundaries of the snowpack.

During the early lifetime of a freshly fallen snowflake, both wind and new snow result in mechanical rounding of grains. Dry snow is characterized by the absence of liquid water. Temperature gradients drive metamorphism in dry snow through the process of vapour diffusion along a vapour density gradient. In *equilibrium growth metamorphism* the sharp edges of new stellar snow crystals are changed into rounded grains due to sublimation from the convex grain surfaces and transport of water vapour to the concave parts (Colbeck, 1982a). *Sintering* between the snow grains forms strong ice bonds (Colbeck, 1997). *Kinetic growth metamorphism* changes the rounded grains into faceted crystals by evaporation and recondensation of water molecules (Colbeck, 1982b).

Wet snow is characterized by a significant amount of liquid water in snow. Liquid water becomes mobile when the irreducible water content is exceeded (about 2-5% by volume; Colbeck, 1982a, 1983). At low liquid-water contents the crystals join together by ice-to-ice contacts and at higher liquid contents there is no intergranular bonding in *wet snow metamorphism* (Colbeck, 1997).

Melting and refreezing cycles create hard, strong ice layers and grain clusters (Colbeck, 1982a). *Firnification* slowly changes snow into glacier ice due to *melt-freeze metamorphism* and the overburden pressure of falling snow (Fig. 2). When snow has changed into ice, the snow density attains a value of 830 kg m^{-3} and the air pores have encapsulated air bubbles (Paterson, 1994).

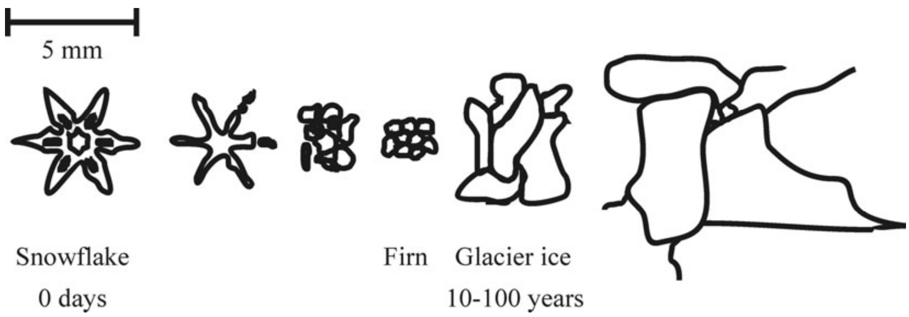

Figure 2. The transition from snow crystals to firn (= snow that has survived one summer) and further to glacier ice (after Hambrey and Alean, 2004). The time scale is dependent on the location (Paterson, 1994).

2.2 Physical properties of snow

2.2.1 Grain size and shape

From the time a snowflake forms until its destruction, it undergoes metamorphism which alters the size and shape of the grain (Fig. 3). Snow-grain size is used to describe the texture and roughness of the snow. One common definition for the grain size is the greatest extension of the grain (usually measured in millimetres) or classification from ‘very fine’ to ‘very coarse’ (Colbeck et al., 1990). Snow-grain size and shape influence the snow density and amount of free water in the snow cover. The determination of snow-grain type and size is crucial for validating snow models and interpreting remote-sensing data (Lesaffre et al., 1998).

Colbeck et al. (1990) collaborated to develop an international classification for seasonal snow on the ground, dividing snow grains into nine different classes based on morphological features with additional information on physical processes and strength (Table 1). Depth hoar is an extreme example of the faceted crystals that are formed in snow when the snow is subjected to a large temperature gradient (Colbeck, 1989). Water vapour moves in the snow cover from the warmer parts to the colder parts. Layers of faceted crystals observed growing immediately below the surface of high-altitude snow covers and polar snow they are explained by a combination of high-temperature gradients associated with the fluctuations of surface temperature and the increased temperature and temperature gradients due to solar radiation input (Colbeck, 1989). Depth hoar formation occurs at temperature gradients of about -10 to –25 °C m^{-1} and snow densities less than 350 kg m^{-3} (Akitaya, 1974; Colbeck, 1982a).

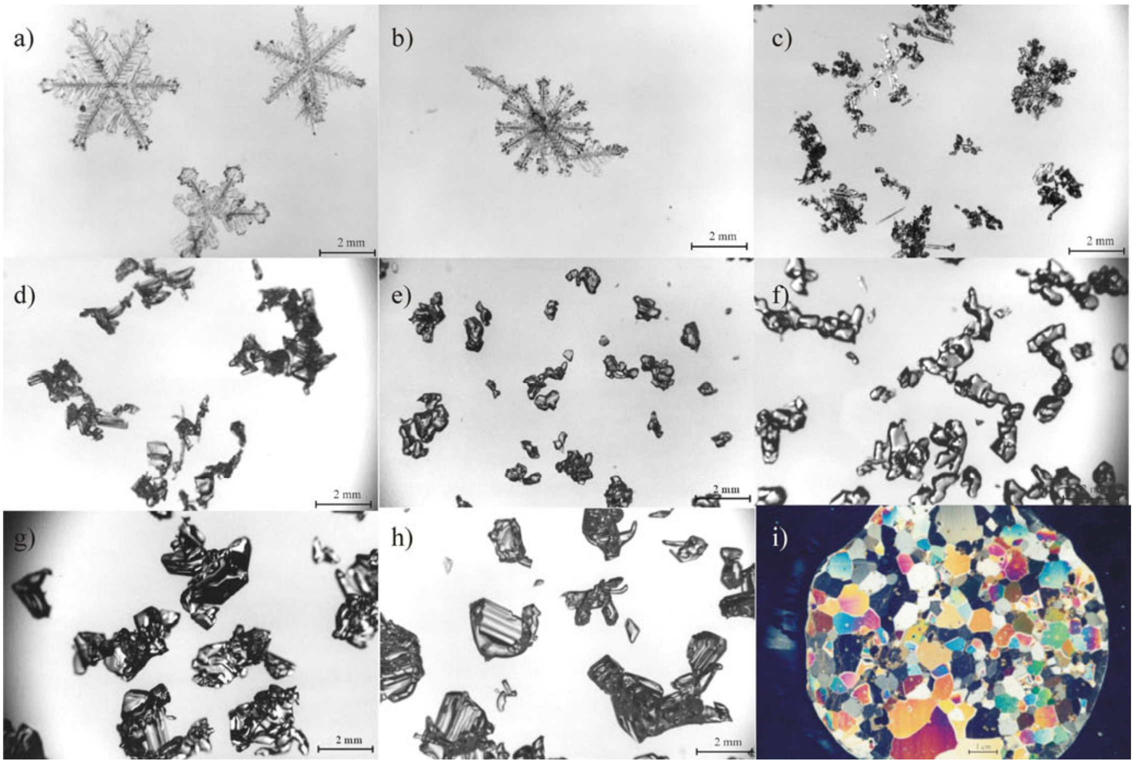

Figure 3. Different snow grains: a) stellar precipitation crystals, b) 12-sided precipitation crystals can sometimes also be seen, c) partly rounded precipitation particles on the ground, d) faceted surface hoar grains, e-f) rounded grains, g-h) cup-shaped depth hoar grains and i) ice crystals from the ice layer found in firn. (Snow-grain photos taken during FINNARP 2000 expedition and firn core sample taken during FINNARP 1999 expedition).

Table 1. Grain shape classification after Colbeck et al. (1990).

Class		Symbol
1	Precipitation particles	+
2	Decomposing and fragmented precipitation particles	⁄
3	Rounded grains (monocrystals)	•
4	Faceted crystals	□
5	Cup-shaped crystals (depth hoar)	∧
6	Wet grains	○
7	Feathery crystals	∨
8	Ice masses	▬
9	Surface deposits and crusts	∨

2.2.2 Density

Densification of snow can be influenced by many different external conditions, particularly in the uppermost portion of the snow cover; e.g. wind-packing, sublimation, melting and freezing may cause a temporary increase or decrease in snow density

(Kojima, 1964). Snow density may vary among different regions, due to local meteorological conditions. The vertical density profile of the uppermost 10-20 m of polar firn is usually described as a function of overburden pressure and local physical quantities such as annual mean temperature, wind speed and accumulation. Higher temperatures enhance grain growth, increasing the densification rate, while mechanical rounding of snow particles by wind action facilitates settling and enhances density in the upper metres; the latter effect is dampened by high accumulation (Craven and Allison, 1998). As snow is deposited at the surface of a snowpack, compaction occurs in two stages (Gray and Morland, 1995). First, there is an initial period of settling when the rate of volume decrease is dominated by thermal processes, reflecting the rapid metamorphism as branched crystals break down (Gray and Morland, 1995). This is followed by a slower densification as pores collapse, which is largely caused by the overburden (Gray and Morland, 1995). Densities increase more slowly at this stage until the intercommunicating air passages become closed off, forming individual bubbles at a density of 830 kg m^{-3} while firn changes to glacier ice (Paterson, 1994). In glacier further densification occurs by compression of the bubbles (Paterson, 1994).

Many of the physical processes occurring in a snowpack are related to snow density. For example, density is an important factor in heat and mass transport; high density favours heat conduction through the ice-grain lattice, while low density favours the processes of diffusion and convection (Colbeck, 1993). The density also influences the mechanical behaviour of snow (Mellor, 1975). Most other snow properties are also related to the density. Snow density affects the effective thermal conductivity and permeability of gases and determines the snow-water equivalent and dielectric properties of snow (Hallikainen and Winebrenner, 1992).

Snow is a granular medium in which a continuum approach is used for length scaling. Thus, continuum quantities such as density are defined for continuum length scales only. Therefore the length scale and variations in grain size are much smaller than we are able to measure with manual density measurements.

2.2.3 Dielectric properties and wetness

The relative dielectric constant of a medium ε is a dimensionless complex number and consists of a real (ε') and an imaginary part (ε''),

$$\varepsilon = \varepsilon' - j\varepsilon'' \qquad (1)$$

where $j = \sqrt{-1}$. The term ε' is usually referred to as the permittivity of the material and ε'' the dielectric loss factor (Ulaby et al., 1986). The real part ε' gives the contrast with respect to free space ($\varepsilon'_{\text{air}} = 1$), whereas the imaginary part ε'' gives the electromagnetic loss of the material. Treatments of the dielectric properties of snow usually result in division of the snow into two groups: a) dry snow, which is a mixture of ice and air and contains no free (liquid) water and b) wet snow, which does contain free water (Ulaby et al., 1986). Electromagnetically, dry snow is a dielectric mixture of ice and air and, therefore, its complex permittivity is governed by the dielectric properties of ice, snow density and ice-particle shape (Hallikainen and Winebrenner, 1992). Since the real part of the permittivity of ice is $\varepsilon'_{\text{ice}} = 3.17$ at frequencies between 10 MHz and 1000 GHz and is practically independent of temperature, the dielectric constant of dry snow is only a function of density $\varepsilon' = \varepsilon'(\rho)$ (Ulaby et al., 1986). Wet snow is a three-component dielectric mixture of ice particles, air and liquid water, and its permittivity is a function of frequency, temperature, volumetric water content, snow density, and the shapes of ice particles and water inclusions (Hallikainen and Winebrenner, 1992). Since the permittivity of water ($\varepsilon'_{\text{water}}$ = 80) is substantially higher than that of ice and air, the dielectric behaviour of wet snow is governed by the volume fraction of water (Hallikainen and Winebrenner, 1992).

Snow wetness or free-water (liquid) water content is typically obtained using calorimetry, dilution or dielectric measurements. Colbeck et al. (1990) gave a general classification of liquid-water content: snow is said to be dry when the percentage of liquid water is 0% by volume, moist at 3%, wet at 3-8%, very wet at 8-15% and saturated slush at >15%. Liquid water is mobile only if the so-called irreducible water content is exceeded and surface forces cannot hold the water against the gravity. The irreducible water content is about 3% and is dependent significantly on snow texture, grain size and grain shape (Colbeck et al., 1990).

2.2.4 Temperature

A snowpack exchange heat with the environment. The heat exchanges either modify the temperature distribution or result in phase changes. If there is no horizontal transfer of heat, the components of the energy budget include net radiation at the snow surface, sensible heat flux, latent heat flux and a term accounting for melting and refreezing. The sources and sinks of the energy budget, along with the radiative and thermal properties of the snow, determine the temperature structure below the surface (Brandt and Warren, 1993). Within an ice sheet, heat is transferred primarily by conduction but other heat-transfer mechanisms can also contribute to varying degrees. Nonconductive processes are limited to the uppermost few metres of snow and can include wind-generated ventilation of the snowpack (wind-pumping), latent-heat transfer by water-vapour migration, convection of air in the pore spaces and solar radiative heating (Brandt and Warren, 1997). In wet snow, internal heat fluxes are controlled by conduction and by latent-heat release due to freezing. Observations of temperature maxima at depths of about 10 cm in cold Antarctic snow during summer were explained by proposing that solar heating is distributed with depth, whereas thermal infrared cooling is localized at the surface (the 'solid-state greenhouse') (Brandt and Warren, 1993).

2.2.5 Stratigraphy

The layered character of snow cover results from a sequence of weather events and metamorphic processes (Fig. 4). The upper layers are largely influenced by solar radiation penetration and diurnal reversals of the temperature gradient (Colbeck, 1991). Hoar layers, crusts and ice layers are examples of layers found in the snowpack. Even if metamorphic processes did not occur, a typical snow cover would be layered just because of varying meteorological conditions (Colbeck, 1991).

Depth hoar and buried surface hoar are usually uniform layers. The development of depth hoar due to vertical vapour transport is dependent not only on the temperature gradient but also on the length of the period it remains near the surface, reflecting the period of nonaccumulation (Mosley-Thompson et al., 1985). Goodwin (1991) observed three distinct seasonal snow surfaces forming in autumn, spring and summer; the autumn and spring wind-glazed ice crusts formed during a hiatus in the snow supply, while the summer surface was characterized by a thin ice layer that formed when maximum radiation melts a thin film over wind-laminated drifting snow. Percolating meltwaters form thick ice layers and lenses. As percolating water reaches a buried wind crust, it is likely to spread along that crust due to its high capillarity and reduced the permeability associated with its small grains and high density (Colbeck, 1991). Snowmelt in Antarctica is limited and occurs occasionally during favourable meteorological conditions; thus the melting phenomena detected may be sensitive to climate change and could be used as climate indicators (Winther, 1993).

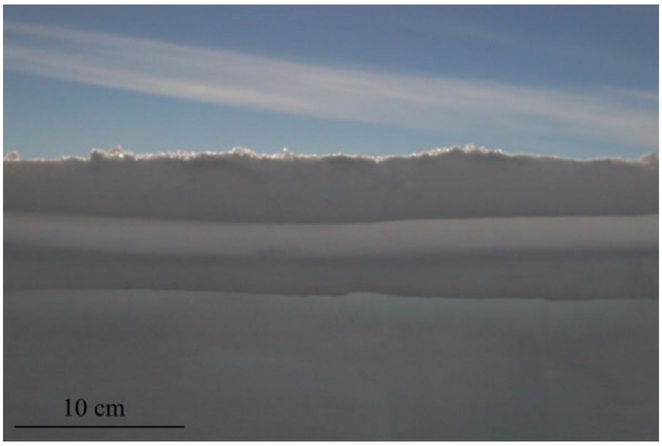

Figure 4. Layered structure of snow cover (photo taken during FINNARP 2000 expedition).

2.3 Snow mass balance

The surface mass balance is the difference between gross accumulation rate (from precipitation, drifting snow deposition, condensation, vapour-to-solid sublimation and superimposed ice formation) and gross ablation rate (from snow deflation, evaporation, sublimation and surface melt runoff) (Giovinetto et al., 1990). The annual accumulation layer is, additionally, subjected to evaporation and melting, especially in the coastal areas. Usually in Antarctica, the meltwater refreezes within the annual layer, which means no loss for the mass balance (Schlosser, 1999) and, except in the coastal zone, melting is small (Van den Broeke et al., 1999).

Snow accumulation (expressed in mm w.e. a^{-1} or kg m^{-2} a^{-1}) is controlled primarily by atmospheric conditions and the elevation and slope of the terrain. A decrease in air temperature usually leads to a decrease in accumulation rates, since colder air can hold less moisture than warmer air, so precipitation amounts tend to decrease, even if the number of precipitation events does not change (Schlosser and Oerter, 2002a).

Precipitated snow is strongly affected by snowdrift, and hence an originally smooth snow cover can be redistributed into a complicated accumulation pattern. The wind redistribution of snow includes erosion of snow cover by the shear force of the wind, transport of blowing snow from exposed sites with low aerodynamic roughness, sublimation of blowing snow in transit and deposition of snow at sites with higher aerodynamic roughness or less exposure to wind (McKay and Gray, 1981). Three modes of drift transport have been distinguished: surface creep, saltation and turbulent suspension (Granberg, 1998). Sublimation from blowing snow constitutes a significant loss of surface snow in dry air.

Intranival air movements are important in snow-drifting, because they aid in the mobilization of drifting snow (Granberg, 1998). They are also important for heat, moisture and gas transfer in the snow cover and, as such, are important for snowpack metamorphism (Granberg, 1998). Wind-pumping driven by air pressure gradients can generate rapid variations in temperature and vapour pressure in snowpack (Granberg, 1998).

Giovinetto and Zwally (2000) showed that the mean accumulation for Antarctica is 159 kg $m^{-2}a^{-1}$ and when taking into account the coastal deflation (the mass loss due to snow removal by wind and sublimation) and ablation (the mass loss due to runoff and evaporation) it is 149 kg $m^{-2}a^{-1}$. The contribution of sublimation in blowing snow to the surface mass balance of the Antarctic ice sheet is significant in the coastal area (0.087 mm w.e. d^{-1}, Gallée, 1998; up to 17 cm w.e. a^{-1}, Bintanja, 1998). Iceberg calving and basal melting are the main ablation processes in the Antarctic ice sheet.

Many authors have studied changes in accumulation rates in various parts of Antarctica. Schlosser and Oerter (2002a) compiled some of these results and concluded that there is no uniform trend found over Antarctica. The West Antarctic Ice Sheet is probably thinning overall, with thickening in the west and thinning in the north. The mass imbalance of the East Antarctic Ice Sheet is likely to be small, but even its sign cannot yet be determined (Rignot and Thomas, 2002).

2.4 Snow chemistry

2.4.1 Oxygen isotope ratio

The stable oxygen isotope ($\delta^{18}O$) ratios of snow are fairly well correlated with the annual mean air temperature at the deposition site, although they are dependent in a complex way on the source and distance to the source of precipitation and on fractionation processes during the transport of moisture to the deposition site of snow (Dansgaard, 1964). The $\delta^{18}O$ value, expressed in ‰, describes the relative deviation of ^{18}O in the precipitation compared with that in Standard Mean Ocean Water (SMOW) (Dansgaard, 1964):

$$\delta_{OX} = \frac{^{18}O/^{16}O(sample) - ^{18}O/^{16}O(SMOW)}{^{18}O/^{16}O(SMOW)} \qquad (2)$$

To use the $\delta^{18}O$ profile for calculating annual accumulation, it is essential that there is precipitation during both the warm and the cold seasons so that the difference in isotopic distribution will be discernable despite losses by wind-scouring and firnification processes (Isaksson and Karlén, 1994a).

Mass exchange by diffusion of water vapour and sintering in the upper layers of the firn smoothes the isotopic profile over a period ranging from months to decades (Johnsen, 1977). Johnsen (1977) showed that this process is no longer important when the density exceeds 550 kg m^{-3}. Schlosser and Oerter (2002b) found that the main part of attenuation of the seasonal signal of $\delta^{18}O$ occurs during the first months of snow metamorphism.

Several natural processes result in isotopic fractionation, or separation of heavier from lighter isotopes, in a sample. Fractionation processes occur at most of the phase changes of water during its atmospheric cycle and can often obscure seasonal or annual variations in the isotopic content of precipitation (Jouzel et al., 1997). The condensed phase (either liquid or solid) is, at equilibrium, isotopically enriched with respect to the vapour phase (Jouzel et al., 1997).

2.4.2 Deposition and sources of ions

The most important sinkage mechanisms for aerosols are precipitation (scavenging processes), dry deposition of particles and adsorption of gases (Fig. 5). The two latter mechanisms result in the snow surface having higher ion contents than in the 'original' precipitation (Gjessing, 1984). On the Antarctic plateau the dry deposition and adsorption of gases are the most important sinkage mechanisms, due to their low precipitation rates, while in the coastal areas scavenging processes predominate (Gjessing, 1984). Dry deposition is the direct deposition of particles from the atmosphere to the snow surface. Aerosols may be directly deposited, whereas gaseous species may also be adsorbed.

In Figure 6 are seen the main origins and sources of soluble impurities found in polar snow. Sea salt is produced by wind action over the open ocean surface (Mulvaney et al., 1993). In atmospheric chemistry studies, either sodium (Na^+) or chloride (Cl^-) is generally chosen as the marine reference element, but magnesium (Mg^{2+}) has also been used (Legrand and Delmas, 1988). HCl was proposed as the major source for excess Cl^- in Antarctica, which is formed by the reaction of excess sulphate with sea-salt particles in the aerosol phase (Legrand and Delmas, 1988). The reaction is more efficient when weather conditions are calm, usually during summer. More than 90% of anthropogenic HCl emissions are confined to the Northern Hemisphere and only a tiny fraction of the emitted amounts of HCl is probably able to reach high southern latitudes (Legrand and Delmas, 1988).

Calcium (Ca^{2+}) and potassium (K^+) have two sources, a marine source as well as continental dust (Legrand and Mayewski, 1997). Ca^{2+} in precipitation is derived from soil dust (primarily as $CaCO_3$), while other Ca^{2+} sources in Antarctica include the marine aerosols (Gjessing, 1984). The crustal source appears to be less important to K^+ than Ca^{2+} (Proposito et al., 2002).

Various natural sources contribute to the natural sulphate (SO_4^{2-}) burden of the atmosphere, in addition to anthropogenic SO_2 emissions (Legrand and Pasteur, 1998). Atmospheric sources of nonsea-salt sulphate ($nssSO_4^{2-}$) and dimethyl sulphide (DMS, CH_3SCH_3) emissions from the marine biota represent a major natural source in the remote marine atmosphere (Andreae and Raemdonck, 1983). In contrast to SO_4^{2-}, the only atmospheric source for methanesulphonic acid (MSA, $CH_3SO_3^-$) is the oxidation of DMS (Legrand and Pasteur, 1998). The MSA to $nssSO_4^{2-}$ ratio has been used in evaluating the contribution of biogenic sources to total SO_4^{2-} over the area and typically shows a summer maximum and a winter minimum. Colder temperatures favour the formation of MSA and could have played a crucial role in controlling the final composition of the high southern latitude atmosphere over the last climatic cycle (Legrand et al., 1992). High MSA concentrations appear to correlate with major ENSO (El Ninõ-Southern Oscillation) events (Legrand and Feniet-Saigne, 1991).

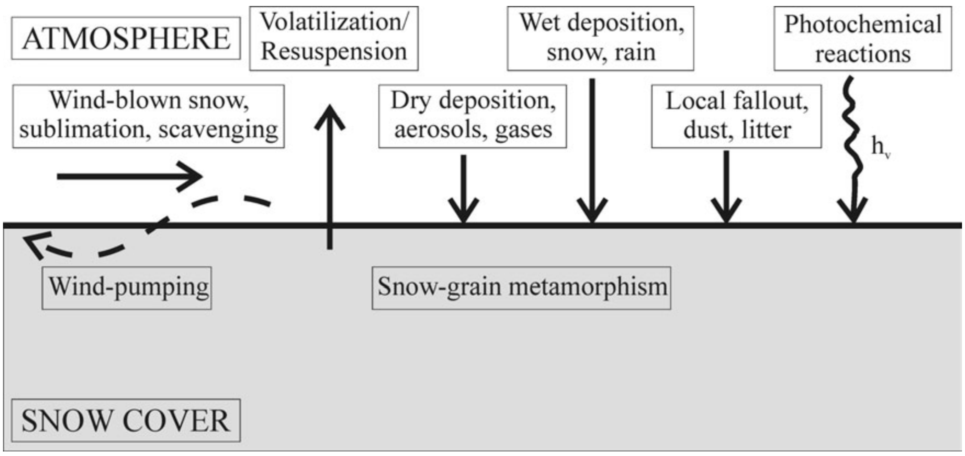

Figure 5. The main physical and chemical processes that influence the chemical composition of cold, dry snow cover (modified after Tranter and Jones, 2001).

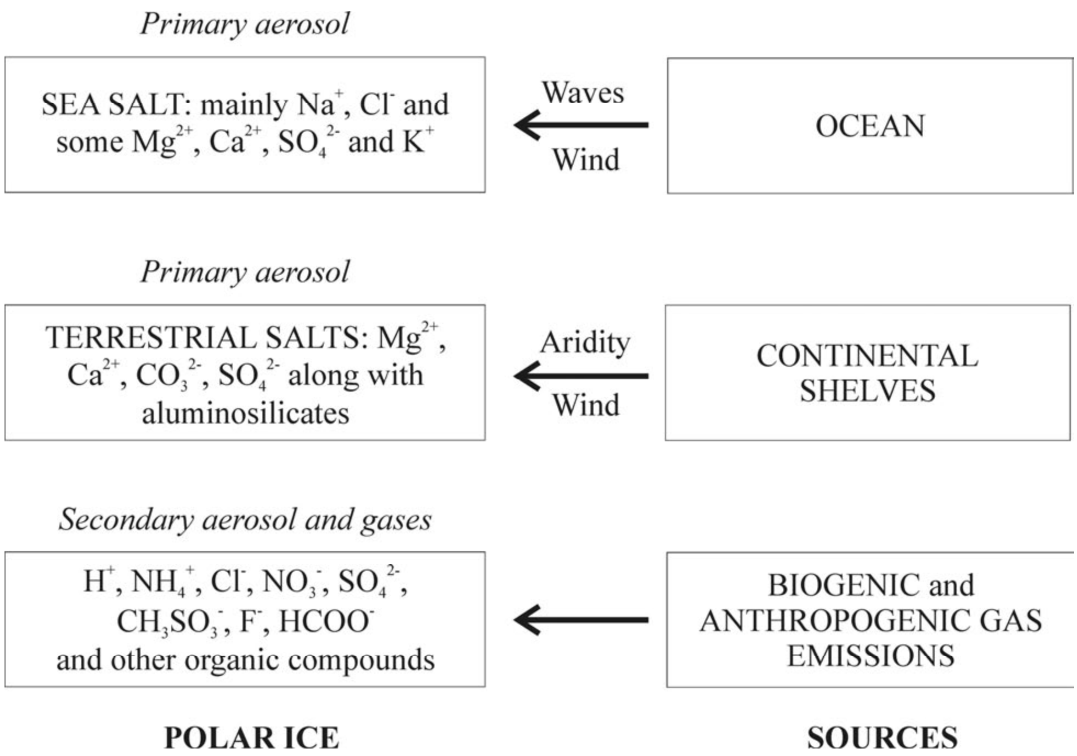

Figure 6. Various soluble impurities trapped in polar snow layers and their corresponding origins and sources (after Legrand and Mayewski, 1997).

Nitrate (NO_3^-) in polar snow is mainly due to the deposition of gaseous nitric acid (HNO_3), an acid that is the end product of the oxidation of various nitrogenous trace gases (NO_x) (Delmas, 1992). Jones et al. (2000) recorded the photochemical production of nitric oxide and nitrogen dioxide (NO and NO_2) in Antarctic snow cover. On the other hand NO_3^- has found to be affected by post-depositional losses at locations of low accumulation (Röthlisberger et al., 2000).

Ammonium (NH_4^+) is mainly released from combustion, bacterial decombustion of plant matter in soils and bacterial decomposition of excreta (Legrand et al., 1998). In the high southern latitudes the primary natural source of NH_4^+ is biogenic oceanic emissions (Legrand et al., 1998) and, to a lesser extent also, animal waste.

2.4.3 Acidity

Snow acts as a reservoir for acids and storage of aerosols. Falling snow collects impurities that accumulate on the snow cover from the dry deposition of gases and atmospheric particles. The acidity of a melted sample is determined by the pH (hydrogen ion potential), defined as the hydrogen ion activity:

$$pH = -\log\left[H^+\right] \tag{3}$$

where $[H^+]$ is the hydrogen ion concentration in an aqueous solution.

The pH value can also be estimated from the ionic balance:

$$pH = -\lg\left[10^{-6}(\sum anions - \sum cations)\right] \tag{4}$$
$$\left[anions\right] = \left[SO_4^{2-}\right] + \left[MSA^-\right] + \left[NO_3^-\right] + \left[Cl^-\right]$$
$$\left[cations\right] = \left[NH_4^+\right] + \left[Na^+\right] + \left[K^+\right] + \left[Ca^{2+}\right] + \left[Mg^{2+}\right]$$

2.4.4 Electrical conductivity

Electrical conductivity (κ; expressed as $\mu S\ cm^{-1}$) measurements of melted samples reflect the conductivity of all ions present in the meltwater (Hammer, 1983). Conductivity shows the ability of an aqueous solution to carry an electrical current and is used especially for melt-water quality control. Dissolved salts in solution carry the current and the conductivity is also dependent on the temperature. It is also partly influenced by the pH and the amount of atmospheric carbon dioxide, which has been dissolved in the water to form ions. The conductivity of snow cover is dependent on the season and is helpful in exact dating (Schlosser, 1999). Conductivity records may reflect individual storm events (higher NaCl), but the signals may be damped by vapour diffusion deeper in the firn (Mosley-Thompson et al., 1985).

3. Methods

The field campaigns were performed in western Dronning Maud Land as a part of the Finnish Antarctic Research Programme (FINNARP) expeditions during the austral summers of 1999/2000, 2000/2001 and 2003/2004. The measurements were conducted along a 350-km transect from the seaward edge of the small Riiser-Larsen ice shelf, along the Ritscherflya ice sheet to Amundsenisen on the plateau (Fig. 7). Snow in the Kvitkuven ice rise and the Högisen ice dome was also investigated. The location of the Finnish Aboa research station and the land routes used determined the measurement sites. Kvitkuven and Högisen were chosen due to lower backscattering observed in the RADARSAT mosaic (Fig. 7). Aboa (73°03'S, 13°24'W) is located on the Basen nunatak at an elevation of 484 m above sea level (a.s.l.). Basen is the northernmost nunatak of the Vestfjella mountain range near the grounding line. Vestfjella and Heimefrontfjella mountain ranges are located approximately parallel to the coast. The measurements were made approximately during a one-month time period. The geographic positions of the snow measurement sites were determined with a hand-held global positioning system device (Garmin, Olathe, KS, USA) with a precision of ±100 m or better.

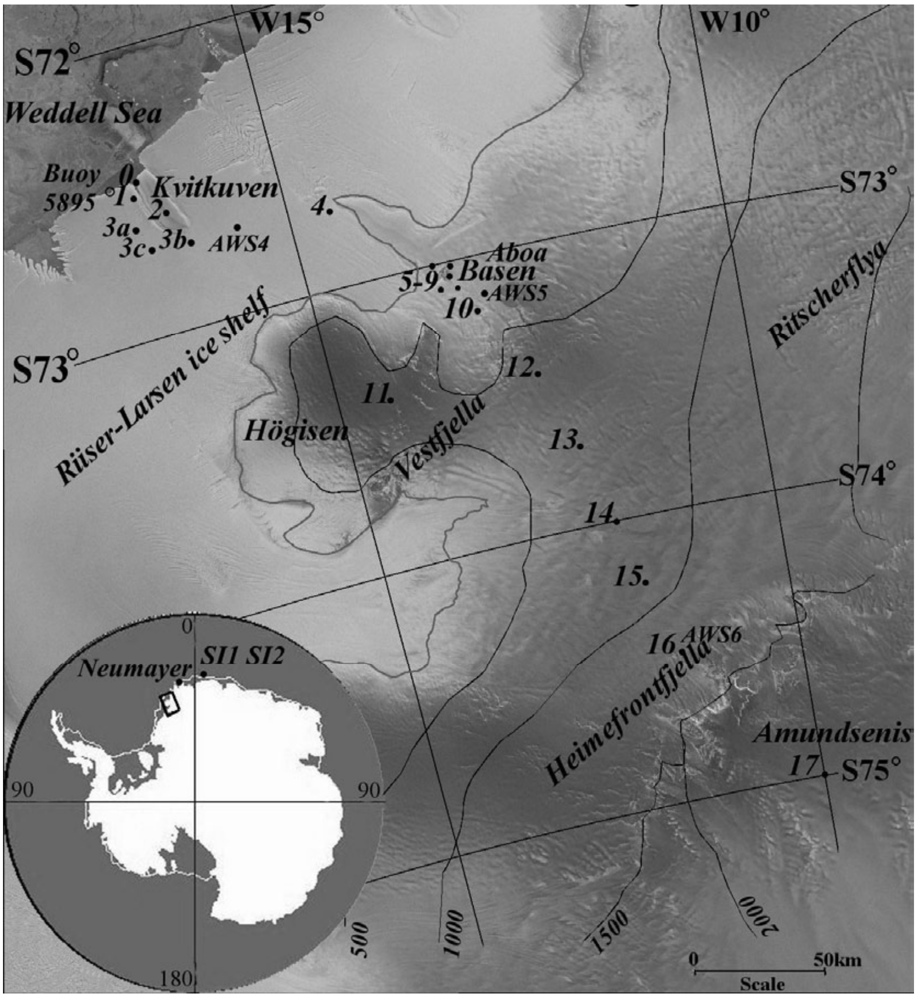

Figure 7. Map of the measurement area in western Dronning Maud Land, showing the locations of snow pit sites 1-17, automatic weather station (AWS) sites and sea ice sites (0, SI1 and SI2). The Finnish Aboa research station (73°03'S, 13°24'W) is located on the Basen nunatak. Locations of Buoy 5895 and the German Neumayer research station have also been marked (article V). Part of the RADARSAT mosaic (Jezek, 1999) was used as background (RADARSAT data © Canadian Space Agency 1997).

Isaksson and Karlén (1994b) have summarized the previous glaciological work done in western Dronning Maud Land before the season 1988/1989. The Norwegian-British-Swedish Antarctic Expedition (1949-1952) started the glaciological research in this area (Schytt, 1958) and the research has continued through decades as mentioned in the article II.

A total of 17 snow pit sites were measured in 1999/2000, 11 sites in 2000/2001 and 10 sites in 2003/2004 on the continent (Table 2). Additionally, there were three sites (SI 1, SI 2 and 0) on the sea ice in 1999/2000 and snow cover on sea ice was measured.

The measurements were conducted in situ in shallow snow pits (1-2 m; Fig. 8) and consisted of profiles at 2-10-cm intervals of visible stratigraphy, temperature, density, grain size and shape, dielectric constant and wetness. The samples were collected for conductivity, pH, $\delta^{18}O$ ratio and chemical analyses (Fig. 9). The temperature profiles were measured using a Pt1000 temperature sensor with ±0.2 °C accuracy. The snow density was measured to an accuracy of ±10 kg m^{-3}, using a cylindrical sampling kit with a volume of 0.5 dm^3 (diameter 6 cm) in 1999/2000 and 0.25 dm^3 (diameter 5 cm) in 2000/2001 and 2003/2004, and a spring balance. In 1999/2000 some missing density data were replaced with values obtained by drilling while installing automated snow sensors at the sites. The error of density measurements σ_ρ was calculated in the following:

$$\sigma_\rho = \sqrt{\left(\sigma_m \frac{\partial \rho}{\partial m}\right)^2 + \left(\sigma_V \frac{\partial \rho}{\partial V}\right)^2} \qquad (5)$$

including density (ρ), mass (m), volume (V), mass error ($\sigma_m = 10^{-3}$ kg) and volume error (σ_V) that includes an estimate for a depression of 0.5 cm in the snow. The density error was estimated to be 3% using the cylindrical sampling kit with a volume of 0.5 dm^3 and 4% using the kit with a volume of 0.25 dm^3.

Snow grains were photographed in the field using a special camera stand (Pihkala and Spring, 1985; Figs. 3 and 10) and classified according to Colbeck et al. (1990). The recorded snow-grain size is the greatest extension of the grain (Colbeck et al., 1990). Here the grain sizes were determined from digital images using ImageJ image-processing software.

The dielectric constant was measured with a TEL 051 dielectric probe (LEAS, Grenoble, France) in 1999/2000 and with the Finnish Snow Fork device in 2000/2001 and 2003/2004 (Sihvola and Tiuri, 1986; Fig. 11). The Snow Fork F9901 sensor was used in 2000/2001 and the FK0101 sensor was also used in 2003/2004. The TEL 051 is a cylindrical probe with a diameter of 6 cm. It measures the real part of the dielectric constant and requires a manually measured density as an input to give the liquid-water content of the snow. The Snow Fork measures both the real and imaginary parts of the dielectric constant to an accuracy of ± 0.02 for the real part and ± 0.002 for the imaginary part. It also provides the density and liquid-water content using an empirical formula with an accuracy of ± 5 kg m^{-3} for density and ± 0.3% for wetness. The accuracies are determined by the manufacturer. The Snow Fork is a fork-type sensor, and near the surface the distance to the interface should be about 5 cm to obtain reliable results; the results of the first 4 cm were omitted. The liquid-water content (wetness) is expressed as a percentage by volume.

The conductivity, pH and $\delta^{18}O$ ratio were analysed from the melted samples. The conductivity was determined with handylab LF 513 T (Schott Glas, Mainz, Germany) in 1999/2000 and 2000/2001, and with a CDM210 conductivity meter (MeterLab, Radiometer Analytical, Lyon, France) in 2003/2004, while the pH was determined with a handylab BlueLine 24 pH meter (Schott Glas).

Figure 8. Snow profiles were measured in the excavated snow pits at 100-200-cm depths (photo taken by H. Granberg during FINNARP 1999 expedition).

Figure 9. Snow samples for chemical analyses were collected using sterile overalls (photo taken during FINNARP 2003 expedition).

The $\delta^{18}O$ ratios were analysed for 1999/2000 and 2000/2001 at the Institute of Geology, Tallinn Technical University, using a Delta-E mass spectrometer (Finnigan-MAT, Bremen, Germany) and for 2003/2004 at the Dating Laboratory, University of Helsinki, using a Delta+XL mass spectrometer connected on-line to GasBench II (Thermofinnigan, Bremen, Germany). The samples were measured against laboratory internal reference waters, which were calibrated on the Vienna Standard Mean Ocean Water/Standard Light Antarctic Precipitation (V-SMOW/SLAP) scale. The reproducibility of the replicate analyses was generally better than ± 0.15‰. The $\delta^{18}O$ and density profiles and snow layering were used to determine the annual layers and annual accumulation.

All snow samples for ion analyses were stored frozen and melted shortly before processing. The anions and cations were analysed simultaneously, using two Dionex-500 ion-chromatography systems at the Finnish Meteorological Institute. The analytical errors were typically 5% of the measured concentrations. In some cases the measured concentrations of certain ions were near their detection limit and the analytical error is larger in these cases (10%). The sample-handling procedure (e.g. melting of the snow samples) also added some error to the measured concentrations.

Figure 10. The special camera stand with light source was used to take snow-grain images (photo taken during FINNARP 2003 expedition).

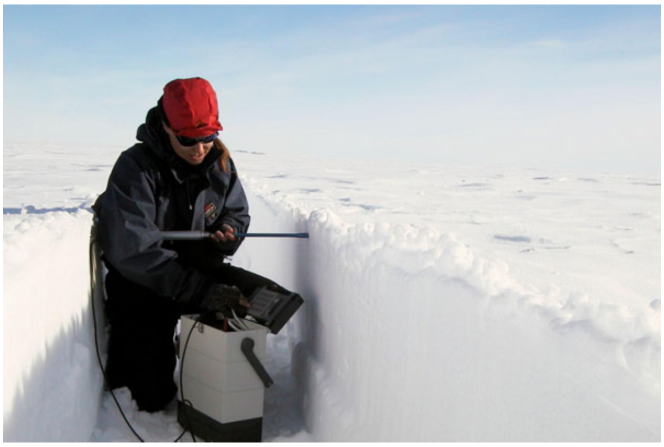

Figure 11. The Finnish Snow Fork device was used to determine the dielectric constant and wetness of the snow cover in 2000/2001 and 2003/2004 (photo taken during FINNARP 2000 expedition).

Table 2. Snow pit site coordinates with elevation, distance from the coast and snow pit depth. SI refers to sea ice and AWS refers to automatic weather station in the area (Reijmer, 2001) and adjacent snow pit sites.

Site	Latitude	Longitude	Approx. elevation m a.s.l.	Distance from coast km	Pit depth 1999/2000 cm	Pit depth 2000/2001 cm	Pit depth 2003/2004 cm
SI1	70°07.4'S	05°23.1'E	0		17		
SI2	70°07.0'S	05°20.7'E	0		20		
0	72°29.3'S	16°31.6'W	0	0	25		
1	72°32.0'S	16°34.0'W	30	3	100	120	152
2	72°36.6'S	16°18.6'W	250	15	150	120	-
3a	72°40.0'S	16°41.9'W	55	20	100	-	-
3b	72°45.0'S	16°00.1'W	60	30	-	130	-
3c	72°45.0'S	16°30.0'W	60	25			112
4	72°45.2'S	14°18.3'W	70	70	200	-	100
5	72°57.9'S	13°34.7'W	270	110	150	115	100
6	73°02.0'S	13°19.5'W	250	120	200	-	-
7	73°03.6'S	13°21.8'W	250	120	100	-	-
8	73°05.3'S	13°20.2'W	240	120	120	-	-
9	73°04.1'S	13°28.2'W	235	120	150	115	-
10	73°12.5'S	13°13.0'W	375	140	150	115	100
11	73°26.3'S	14°26.7'W	990	130	200	150	-
12	73°27.4'S	12°33.3'W	905	170	150	100	100
13	73°43.0'S	12°18.6'W	930	195	110	-	-
14	74°00.8'S	12°01.1'W	980	230	150	110	100
15	74°14.0'S	11°48.0'W	1000	250	130	-	-
16	74°28.7'S	11°33.1'W	1100	275	220	110	-
17	74°59.9'S	10°00.5'W	2550	355	160	105	-
AWS4	72°45.1'S	15°30.0'W	60	45	-	-	4 x 100
AWS5	73°06.2'S	13°09.8'W	370	130	-	-	4 x 100
AWS6	74°29.0'S	11°31.2'W	1100	275	-	-	4 x 100

4. Summary of articles

This thesis consists of four articles published and one submitted in 2002-2006. The articles present the physical and chemical stratigraphic results from the upper 1-2 m of the snow pits and evaluate the spatial variations and seasonal signals of the parameters in western Dronning Maud Land. The annual accumulation rates in the measurement area were calculated. In addition the meteorological conditions near the Aboa station were reported.

4.1 Article I

Kärkäs, E., H.B. Granberg, C. Lavoie, K. Kanto, K. Rasmus and M. Leppäranta. 2002. Physical properties of the seasonal snow cover in Dronning Maud Land, East-Antarctica. *Annals of Glaciology* **34**, 89-94.

In this study the physical properties of the most recent annual snow accumulation were investigated along a 350-km transect from the sea ice to the polar plateau in Dronning Maud Land during the austral summer of 1999/2000 for the presence of spatial and temporal variations. The optical measurements were included and studied together with the other properties. Three of the measurement sites were in very remote locations (Kvitkuven ice rise, Högisen ice dome and Amundsenisen on the plateau).

The results showed widespread spatial variations in snow property profiles and layering, and five principal snow zones with different characteristics in the area were suggested. The seaward edge of the ice shelf was clearly distinguished from the rest of the ice shelf and the local topographic highs were distinct from the nearby ice sheet. The study revealed that the distance from the sea and moisture source affect the snow properties in this coastal location. The small-scale variations were influenced especially by local topography, due to the reduced speed of the katabatic winds.

4.2 Article II

Kärkäs, E., T. Martma and E. Sonninen. 2005. Surface snow properties and stratigraphy during the austral summer in western Dronning Maud Land, Antarctica. *Polar Research* **24**(1-2), 55-67.

This paper compared the physical properties of the surface snow and annual accumulation rates measured in shallow snow pits (1-2 m) during three austral summers (1999/2000, 2000/2001 and 2003/2004) in coastal Dronning Maud Land along a 350-km transect from the ice edge to the plateau. The $\delta^{18}O$ ratios, stratigraphies and annual accumulation rates were examined in detail. A large number of snow-grain images were analysed to determine the snow-grain sizes.

The results reveal wide spatial and temporal variations in the measurement area that are the key to understanding the results obtained from deep ice cores and remote sensing applications. Some measured quantities remained fairly constant with increasing distance from the ice edge to the polar plateau, while the elevation varied from 30 to 2550 m a.s.l. Although some of the properties showed no visible trends from the coast to inland regions, they varied widely at individual sites, reflecting the great importance of these spatial studies. The $\delta^{18}O$ profiles were the most effective tools available for dating the annual accumulation layers. Our results confirmed that conductivity, grain size, $\delta^{18}O$ ratio and accumulation rate clearly decreased with increasing distance inland from the ice edge. The snow temperature correlated most favourably with the surface elevation, which in Antarctica increases with increasing distance from the ice edge. The depth hoar layers, when they occurred, were usually found below the thin, hard ice crust and the Snow Fork device was able to detect these low-density layers. The frequency distribution of the snow-grain sizes was skewed to the right, with the mean grain size varying seasonally between 1.5 and 1.8 mm.

4.3 Article III

Kärkäs, E., K. Teinilä, A. Virkkula and M. Aurela. 2005. Spatial variations of surface snow chemistry during two austral summers in western Dronning Maud Land, Antarctica. *Atmospheric Environment* **39**, 1405-1416.

In this paper the glaciochemical properties of the most recent precipitation, including the topmost 5 cm, of surface snow were measured and their spatial variations investigated during two austral summers in 1999/2000 and 2000/2001, along a measurement transect from the sea ice to the plateau in western Dronning Maud Land. The nss fractions, enrichment factors and ionic balances were calculated and detailed statistical analyses conducted. The ionic concentrations were combined with aerosol measurements performed simultaneously at the Aboa station to determine the fraction of dry deposition in the ionic concentrations.

The results showed that in the summer surface snow the sea-salt components, MSA and $nssSO_4^{2-}$ decreased exponentially with increasing distance from the ice edge, with rates of decrease of 48 - 64% per 100 km and correlation coefficients of –0.77 to –0.89. There was no trend for NO_3^- in the surface snow. Spatial variations within single sites were large, most possible due to wind redistribution. Crustal sources were visible near the nunataks. The results revealed that usually less than 10% of the deposition originated from dry deposition near the Aboa station.

4.4 Article IV

Kanto, E., K. Teinilä, H. Timonen and E. Sonninen. Stratigraphy and spatial variations of snow chemistry in western Dronning Maud Land, Antarctica. Submitted to *Nordic Hydrology*.

In this study the ionic concentrations of the most recent annual accumulation were analysed and investigated together with snow stratigraphy in coastal Dronning Maud Land in 2003/2004. The measurement area was located on the ice shelf and on the ice sheet between two mountain ranges. The high-resolution samples were collected in 19 snow pits. A total of four adjacent pits at each of three AWS locations were measured to determine the local spatial variation. Here we combined the chemical, isotopic and physical stratigraphic profiles to investigate their seasonal and spatial variations. The nss fractions and ionic balances were also calculated.

The results showed that the sea-salt components predominated in the coastal region and that the widest spatial and temporal variations with the most pronounced exponential decreases occurred at progressive distances inland from the ice edge. The ionic concentrations of sea salts at the site on the seaward edge of the ice shelf were anomalous high compared with those at other sites. For the less studied oxalate (Ox^{2-}) the ocean is apparently a major source and its concentrations also clearly decreased with increasing distance from the ice edge. The NO_3^- concentration increased linearly with distance from the coast, in contrast to our earlier results. Seasonal sea-salt variations are more visible behind the grounding line, indicating the interference of strong summer storms in the flat ice shelf area. The concentrations and seasonal cycles varied widely among pits at single sites mainly due to the redistribution of snow by wind and sastrugi topography, thus revealing the necessity for further detailed investigations of spatial variations. The mean NO_3^- values of the annual layers increased linearly with increasing distance from the ice edge and the highest concentrations were detected in association with the depth hoar layers. Distinct summer surface peaks were observed with the nitrogen- and sulphur-containing compounds and were good dating tools for use in the measurement area. About two thirds of the sea-salt peaks occurred during winter showing the uncertainty of the use of sea salts for accurate dating.

4.5 Article V

Kärkäs, E. 2004. Meteorological conditions of the Basen nunatak in western Dronning Maud Land, Antarctica, during the years 1989-2001. *Geophysica* **40**(1-2), 39-52.

In this paper, the meteorological data from the Aboa station covering the time period 1989-2001 and showing the special microclimate conditions in the nunatak environment, were statistically analysed. In 2001 this old AWS broke down and a new station was set up during the 2002/2003 season at a different location. This study comprises the whole existing dataset from the previous weather station. The monthly and annual mean values as well as the recorded extremes were reported. The data were compared with some other AWSs in Dronning Maud Land.

The data showed that the wind directions were locally redistributed due to the nunatak and that the nunatak also affected the summer air temperature values. A total of 77% of the observed wind speeds were ≤ 10 m s^{-1}; highest wind speeds were recorded during June and August and were connected with cyclonic storms. The mean annual air temperature at Aboa was –15 ˚C; the air temperature and wind speed displayed annual cycles and a semiannual oscillation was seen in the air pressure data. The German Neumayer station was also a good reference for the Aboa area, even though it is located about 330 km away on the ice shelf.

4.6 Author's contribution

The author's own contribution to each publication mentioned above is shown in Table 3, classified into three categories: theory, technique (includes sample collection, preparation and analytical work) and writing (includes data analysis and writing of the articles).

Table 3. Percentages of author's own contribution to each publication:

Paper	Theory	Technique	Writing
I	>70	70	70
II	100	50-70	100
III	50-70	50-70	70
IV	>70	50-70	>70
V	100	-	100

In I, the first author was responsible for collecting and processing all data, except the radiation measurements and the δ^{18}O analysis, and for most of the writing except the radiation measurements and part of the Introduction; the other authors were responsible for Figures 1 and 4. The general background behind the study was based on the Master's thesis of the first author.

In II, the first author performed the field measurements and collected the snow samples for the $\delta^{18}O$ analyses, and was responsible for the data analysis, interpretation, all figures and the writing. The other authors were responsible for the $\delta^{18}O$ analysis.

In III, the first author collected the snow samples and was responsible for the data processing, figures and most of the writing. The other authors were responsible for the chemical analyses, for collecting and analysing the filter samples and suggested that calculating the rate of decrease would aid in interpretation of the data.

In IV, the first author collected the snow samples and was responsible for the snow stratigraphy studies, most of the interpretation, writing, all calculations and the figures. The other authors were responsible for the chemical (including $\delta^{18}O$) analyses.

In V, the author was not involved in data collecting, but received the raw meteorological data from the Finnish Institute of Marine Research.

5. Synthesis of results

The variation in all measured properties was wide at single sites, mostly due to the redistribution of snow by winds and sastrugi topography, thus revealing the importance of spatially representative measurements. Detailed measurements of the spatial variations in snow properties and in the annual accumulation of snow on the ice sheet are needed to obtain spatially representative results and to further use them in interpreting ice cores and satellite images.

In the coastal area of western Dronning Maud Land with a gentle surface slope, the distance from the coast is more important factor controlling the variations in snow properties than the surface elevation. Physical properties that are strongly affected by the moisture source (conductivity, grain size, $\delta^{18}O$ ratio and accumulation rate) clearly decreased continuously from the ice edge to the polar plateau. Also the ionic concentrations of sea salts (Na^+, Cl^- and Mg^{2+}), MSA and $nssSO_4^{2-}$ tended to decrease from the ice edge to the inland regions. The concentrations of sea-salt ions of the whole accumulation layer showed more rapid decrease from the ice shelf edge to inland regions than those found earlier in the same area only for summer surface snow. The opposite was found for the MSA and $nssSO_4^{2-}$ concentrations with more rapid decrease of the summer surface values than those found for the whole accumulation year. The NO_3^- concentration showed inconsistent results; while investigating only the summer surface in 1999/2000 and 2000/2001 there was no visible trend from the coast to inland but in 2003/2004 the NO_3^- concentration of accumulation layer increased from the coast to the inland regions. The post-depositional processes might change the NO_3^- concentrations in the snow cover and also the clear seasonality might cause these different results. During the austral summer, the nunataks locally affected on the ionic concentrations of $nssK^+$ and $nssCa^{2+}$ and the pH values.

Many of the measured ionic profiles showed seasonal signals and the chemical records are generally used for dating snow and ice layers. Before interpreting the past, the composition and the origin of the present chemical impurities deposited in polar snow have to be understood. The scavenging processes predominate in the area; usually less than 10% of the total deposition of surface snow was dry deposition near the Aboa station, with the exception of Ca^{2+}. The observed seasonality in the chemical profiles was most pronounced in the nitrogen (NO_3^- and NH_4^+)- and sulphur-containing ($nssSO_4^{2-}$ and MSA) compounds. The $nssSO_4^{2-}$ and $\delta^{18}O$ profiles were the most effective dating tools. Based on measurements performed during three different seasons and the $\delta^{18}O$ profiles the mean snow accumulation on the ice shelf was 312 ± 28 mm w.e., on the ice sheet between two coastal mountain ranges 215 ± 43 mm w.e. and on the plateau 92 ± 25 mm w.e.

The stratigraphy in the coastal area was very diverse and several ice layers and depth hoar layers were found within a single accumulation year. The depth hoar layers were associated with low dielectric constant values. The highest concentration of NO_3^- was detected in association with those low-density, coarse-grained depth hoar layers that usually form under thin ice crust during the hiatus in accumulation in the austral summer or that represent buried surface hoar. The ice layers were thick near the coast and consisted mostly of thin crusts behind the grounding line.

Based on the results the snow properties varied among the ice shelf, coastal region and polar plateau. Heimefrontfjella forms a steep step in the landscape and causes changes in the gradients of the properties studied in the direction of the plateau. In addition to these three snow zones two subgroups could be distinguished: local topographic highs and the edge of the ice shelf. Snow was less densely packed and had a smooth surface due to the reduced effect of wind on the local topographic highs. The seaward edge of the ice shelf was an abnormal site that clearly influenced snow properties (high conductivity and ionic concentrations) and summer melting in the vicinity of sea.

From the visual interpretation of the RADARSAT mosaic a significantly lower intensity of backscattered signal (appearing darker) was observed for the local topographic highs, Kvitkuven ice rise and Högisen ice dome. Both sites had visible smoother surface, lower dielectric constant and snow density values than nearby valleys due to lighter wind action. There were no thick ice layers indicating melting-refreezing on the highs. Heterogeneous snowpacks appear brighter than homogeneous ones. That was also observed at the most coastal site (1) where bigger snow grains, higher dielectric constant values and thick ice layers caused visible stronger backscattering compared to the rest of the measurement area. Backscattering over dry snow is dependent on snow density, grain size, free liquid water content and stratification. The observed snow properties could be used as the validation data in remote sensing applications.

The measured wind speed values from Aboa verify that the snow-drifting could cause notable amount of removal and sublimation from the snow surface and has an effect on the surface mass balance. The maximum air temperatures show that there occurs melting and melt water refreezing in snow cover during summer near the grounding line. Even though the effect of the Basen nunatak could be seen in the AWS data (especially air temperature, wind speed and direction) from Aboa, it could be used to describe the general meteorological conditions in the measurement area and the results correlated well with the data obtained from the German Neumayer station.

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